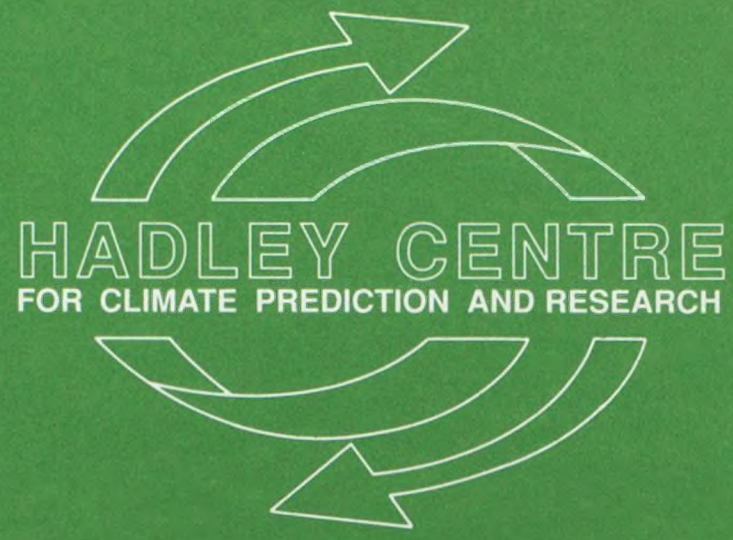


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A. J. Dolman and D. Gregory

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Hadley Centre
Meteorological Office
London Road
Bracknell
Berkshire RG12 2SY

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# THE PARAMETERIZATION OF RAINFALL INTERCEPTION IN GCMs

by

A J DOLMAN

Institute of Hydrology, Wallingford

and

D GREGORY

Hadley Centre

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Hadley Centre for Climate Prediction and Research
Meteorological Office
London Road
Bracknell
Berkshire RG12 2SY
U.K.

*Abstract*

Experiments with a one dimensional version of the UK Met Office 11-layer GCM are used to derive a simple calibrated subgrid parameterization of rainfall interception by Amazonian rainforest. Two interception parameterizations, both incorporating subgrid variability of precipitation within a GCM grid box are presented and tested and compared with observations made at a single location. The 1-D model is a fully interactive atmospheric column and incorporates all the GCM physics but is forced at its boundaries by climatic data. The interception losses appear to be sensitive to the fractional area in the GCM grid box which is covered by the rain and the precise formulation of the interception process. The implications for the parameterization of the wet area in a GCM grid box are discussed and suggestions are made as how to parameterize the fraction of wetted area in a GCM grid box.

*Introduction*

Experiments with general circulation models (GCMs) have demonstrated that the atmospheric circulation and rainfall patterns over the continents may be influenced by the interaction of the land surface and the lowest layers of the atmosphere (Rowntree, 1988, Mintz, 1984). Deforestation experiments with GCMs have shown that substantial changes in atmospheric circulation and rainfall may result from changing the tropical rainforest of the Amazon into pasture or savanna (e.g. Lean and Warrilow, 1989, Dickinson, 1989, Shukla et al, 1990). The results of these experiments generally suggest a weakening in intensity of the hydrological cycle if large areas of forests are cleared. Since about 50% of the rainfall within the Amazon basin is derived from local evaporation (Shuttleworth, et al., 1984), the prediction of the likely climatic effects of deforestation in the Amazon will ultimately depend on the realism of the particular land surface parameterization scheme used (e.g. Henderson-Sellers, 1987).

Recently a number of land surface parameterization schemes have been proposed which incorporate a considerable degree of complexity in their description of the land surface-atmosphere interaction (Sellers et al., 1984, Dickinson, 1984, Noilhan and Planton, 1989). These models are usually designed and calibrated using observations from single sites rather than with area average data (e.g. Sellers et al., 1989). They are nevertheless forced with grid area averaged output variables, such as rainfall and humidity, from the atmospheric circulation model to which they are coupled.

This may lead to problems in the parameterization of spatially heterogeneous, sub-grid physical processes. For instance, rainfall interception is known to be sensitive to the temporal structure of rainfall (Lloyd, 1990). In the case of the Amazon rainforest, where convective rainfall is the dominant rainfall producing mechanism, the grid rainfall rate is a composite of several storms, which may vary in surface coverage, duration and intensity. GCMs at present do not include parameterizations of this variation and tend to overestimate interception loss and underestimate dry canopy evaporation (Lean and Warrilow, 1989, Abramopoulos et al., 1988) when the results are compared with observations (eg Shuttleworth, 1988b).

Pitman et al (1990) report an experiment with a stand alone version of the Biosphere Atmosphere Transfer Scheme (BATS, Dickinson et al, 1986), in which a new rainfall interception parameterization proposed by Shuttleworth (1988b) is used to describe the subgrid variability

of rainfall in a GCM grid. They report a substantial sensitivity of the runoff-evaporation regime of tropical rainforest to the details of this subgrid parameterization. However, the results of their study are of limited value as the atmospheric forcing is prescribed rather than calculated as in a GCM.

This paper describes experiments with a 1-D version of the UK Met Office 11-layer GCM exploring the sensitivity of the predicted Amazonian climate to the description of the wet canopy evaporation from the tropical rainforest. The 1-D model represents a single large area, roughly the size of a GCM gridsquare, and includes all the physics routines of the UK Met Office GCM (Slingo, 1985). At the boundaries horizontal advection of heat, moisture and momentum is prescribed for a specific location using data from Oort (1983). The use of a 1-D model in the context of land surface atmosphere modelling is relatively novel (Warrilow et al, 1986, Koster and Eagleson, 1990). It allows investigation of the interaction of the land surface with the atmosphere and is therefore particularly useful in the development of land surface parameterizations. Results are compared with observations obtained in the Anglo-Brazilian field experiment in the Reserva Ducke near Manaus, Amazonia (Shuttleworth, 1988b).

*1-D model*

The 1-D model was developed by Warrilow et al. (1986) from the UK Meteorological Office 11-layer GCM (Slingo, 1985) used for climate research. Sigma co-ordinates (pressure over its surface value) are used in the vertical and all physical processes included in the GCM are modelled: interactive radiation with four cloud types (low, medium, high and convective), a mass flux convection scheme, large scale precipitation, boundary layer and surface turbulent exchange (all described in Slingo, 1985), together with a land surface scheme, described briefly in the next section. The model is integrated forward with a timestep of 5 minutes with the radiation scheme being called every 2 hours.

The main distinguishing feature of the 1-D model from other similar models is its treatment of heat and moisture advection. Warrilow et al (1986) provide a detailed description of the formulation and its derivation. A brief description is given below. The model is assumed to represent a circular area with radius $\Delta n$. The local rate of change of a quantity $X$ averaged over the area due to large scale advection can be written as

$$\frac{\Delta X}{\Delta t_d} = \frac{|V_n|\,(X_r - X)}{\Delta n} - \frac{\omega \Delta X}{\Delta p} \qquad (1)$$

Where $|V_n|$ is the absolute value of the horizontal velocity in the direction of the positive gradient of $X$ across the area, $\omega$ is the vertical p-velocity, p is the pressure and $X_r$ a reference value of $X$ outside the area under consideration. Knowledge of $V_n$, w and $X_r$ is required if the forcing due to the large scale advection is to be specified. A Gaussian probability distribution for these quantities is assumed with means and standard deviation being taken from the global data set compiled by Oort (1983). To specify the moisture reference profile dew point depression (D) is used. Due to the lack of available data the standard deviation of D is linearly related to that of temperature. Because the mean values of $X_r$ relate to the centre of the area under study, while the reference profile is required at the edge , a correction is applied to account for the gradient of a quantity ($\delta X$) across the area. The velocity component parallel to the positive gradient of $X$ ($V_p$) across the area is also specified for use in the boundary layer scheme.

Data used in the advective forcing scheme is shown in Table 1a and 1b for January and July for the Central Amazon. A sinusoidal interpolation procedure is used to derive values for other months than these. A random number generator is used to specify the exact forcing on any one day, from the assumed Gaussian distributions described above. Values are interpolated from one day to the next to prevent large step changes in the forcing applied to the model. The only tuning in the forcing representation is in the exact relationship between the standard deviation of dew point depression and temperature which is chosen to give the correct mean rainfall for a given area. Initial model results indicated a lack of seasonal variation in the rainfall over the Amazon. Vertical velocities from the Oort data were modified to correct this according to the balance of observed precipitation and evaporation. The revised figures are included in Table 1.

*Land surface scheme*

The land surface scheme presently used in the GCM is depicted in Fig. 1. Surface temperature is calculated using a four layer soil temperature model. This model was designed to produce a reasonable surface temperature response to radiative forcing in the frequencies from 1/2 a day to a year (Warrilow et al., 1986). It is important to stress that the soil-vegetation

layer in this representation is treated as a single solid body with a single surface temperature, equal to the temperature of the top layer, which represents the temperature of the soil and the vegetation. No effort is made to calculate different surface temperatures for soil and vegetation as in some of the more complex recent land surface schemes. The single value of surface temperature is used in both radiative and turbulent exchange calculations. The main parameters used in this scheme are summarized in table 2.

The fluxes of latent and sensible heat are calculated from

$$F = -\delta X \mid V_{11} \mid C_x \qquad (2)$$

where F is either the sensible or latent heat flux, $\delta X$ the change in temperature or humidity from the surface to the lowest atmospheric level and $C_x$ a bulk transfer coefficient, $V_{11}$ is the horizontal wind velocity at the lowest model level. $C_x$ is a complex function of height, atmospheric stability (as expressed by the bulk Richardson number) and surface roughness. In the case of evaporation from vegetated land surfaces the latent heat flux from the soil-vegetation surface is calculated from

$$E = \frac{\delta q}{(r_a + r_s)} \qquad (3)$$

with $r_a$ an aerodynamic resistance, equal to $1/(C_v \mid V_{11} \mid)$ and $r_s$ the surface resistance to water vapour transfer, $\delta q$ is the change in specific humidity from the saturated surface to the lowest model layer. Equation (3) calculates a potential evaporation rate ($E_p$) which is adjusted to the actual rate by reducing the 'unstressed' surface resistance, $r_{su}$ according to:

$$r_s = \frac{r_{su}}{\beta} \qquad (4)$$

where ß is a reduction factor defined by the minimum of 1 and $(\Theta-\Theta_w)/(\Theta_c-\Theta_w)$, and $\Theta$ is the soil moisture content and the subscripts c and w refer to a critical level of soil moisture, and the wilting point respectively. The unstressed surface resistance is set to 115 s $m^{-1}$ in accordance with the results of Dolman et al. (1991). This formulation is somewhat different to that used in the full GCM where the reduction factor ß works directly upon the evaporation, the actual amount being $\beta E_p$. However, since the physiological response of the vegetation to increasing soil moisture deficits is more realistically expressed as an increase in surface resistance, Equations (3) and (4) are employed.

Evaporation from a wet canopy is calculated from equation (3) with the surface resistance set to zero. The canopy store is filled by either large scale or convective precipitation and condensation, and emptied by throughfall and evaporation. Throughfall, (T) is modelled assuming a dependence on the depth of water on the canopy, C, the canopy storage capacity S and the rainfall rate P

$$T = (\frac{C}{S}) P \qquad (5)$$

Note that this formulation implies throughfall during the wetting up phase of the canopy. In the case of a not completely filled canopy store a fraction 1-C/S of the grid area is assumed to be freely transpiring and a fraction C/S assumed to evaporate intercepted rainfall.

*Experiments with standard canopy scheme*

Previous work on forest evaporation has emphasized the need for a separate description of wet and dry canopy evaporation (e.g. Shuttleworth, 1989), an observation which applies equally well to the Amazonian tropical rainforest. In a GCM grid area however, a single set of meteorological variables has to describe both dry and wet conditions and particular problems arise with respect to the averaging techniques used to calculate the dry and wet canopy contributions.

A series of 1-D model experiments was designed to compare the results obtained using the present land surface description of the GCM with results obtained from a locally calibrated model (Shuttleworth, 1988b). Where references in this paper are made to observations of Shuttleworth (1988b), these refer to the synthesised secondary data of Shuttleworth (1988b). The results of a 1-D model experiment, in which the above description (table 2) of the rainforest was used (A1), is compared with the results from Shuttleworth (1988b) in table 3. The 1-D model was integrated for 1 year in these experiments.

The modelled and observed seasonal distribution of rainfall is shown in Figure 2. The seasonal distribution is obtained from a 15 year record of rainfall (1971-1985) in the Reserva Ducke reserve. The agreement between observed and modelled precipitation is reasonable, with rainfall underestimated in the first half of the year (by 15 to 20 %) and overestimated in the second half (by 10 to 15%). Differences between model and observed values can be a result of errors in the response of the convection scheme, the random nature of the forcing or the

particular climate forcing data. Within the present context, the agreement is sufficiently good to proceed with the testing of the interception parameterizations. The model reproduces the results of the GCM experiments, which used essentially the same parameterization (Lean and Warrilow, 1989), with the interception loss accounting for 60% of the total evaporation. The partitioning of the total evaporation between interception (wet canopy evaporation) and transpiration (dry canopy evaporation) is shown in Figure 3. Lean and Warrilow found wet canopy evaporation to account for 71% of total evaporation which was overestimated by 15%, compared with observations. The 1-D model simulations underestimate total evaporation by about 10%, but this difference is probably insignificant when interannual variation in precipitation and evaporation are taken into account. The model uses about 90% of the available energy for evaporation. This agrees with the results of Shuttleworth (1988b). Around 50% of the annual rainfall is returned to the atmosphere by the process of evaporation, in accordance with observations. It is worth noting that in the 1-D model experiment the value of the saturation storage capacity was reduced from 2.5 (Lean and Warrilow, 1989) to 0.7 mm. This did not produce a substantially different interception loss, the percentage interception loss for both cases was 27% and 26% respectively, when the model was integrated for one month. Shuttleworth (1988b) and Lloyd et al. (1988) however, report considerable sensitivity of the modelled interception loss to changes in the value of the saturation storage capacity. The reduction in storage capacity in the 1-D experiment is apparently not sufficient to allow the canopy to become more saturated and to start producing throughfall by canopy drip.

*Experiments with revised canopy schemes*

The overestimation of interception loss is likely to be caused by the neglect of spatial variability in the convective rainfall pattern over a single GCM grid. The time distribution of grid averaged rainfall can be compared directly to the observed time distribution at Manaus as this comparison involves only the mean quantities and does not involve higher order moments (Figure 4). The modelled distribution agrees very well with the observations over a 2 year period at the Reserva Ducke site (Lloyd, 1990). The agreement suggest that the 1-D model correctly reproduces both the average forcing and the dominant convective nature of rainfall in the Central Amazon basin. However, to allow for the spatial variation in this rainfall and for the fact that at any one time the GCM grid is unlikely to be fully covered by a single storm, it is necessary to take these factors into account with the canopy interception formulation.

Shuttleworth (1988a) presents such a parameterization. The local rainfall rate in the rain-covered part of the grid is assumed to follow a negative exponential probability distribution

$$f(P_l) = \left(\frac{\mu}{P}\right) \exp\left(\frac{-\mu P_l}{P}\right) \qquad (6)$$

with P and $P_l$ the model's grid average rainfall rate and the local rainfall rate, and $\mu$ the proportion of the grid covered with rain. Within the area the local throughfall is taken to be:

$$T_l = \begin{cases} P_l - C_m & \text{if } P_l \Delta t > S - C \\ 0 & \text{if } P_l \Delta t < S - C \end{cases} \qquad (7a)$$

where $C_m$=(S-C)/$\Delta$t and $\Delta$t is the timestep. It can be seen that no throughfall ocuurs locally untill the canopy is saturated. It should be noted that this description differs from that used by Warrilow et al (1986) in that respect. Integrating over the area where rain occurs and averaging over the gridbox Shuttleworth (1988a) demonstrated that:

$$T = P \exp\left(\frac{-\mu C_m}{P}\right) \qquad (7b)$$

Recently an analogous scheme, based on the original Warrilow et al (1986) assumption of the dependence of throughfall on rainfall rate (Equation 5), has been developed and included in the new UK Meteorological Office Unified Modelling system (Gregory and Smith, 1990). :

$$T_l = \begin{cases} P_l - C_m & \text{if } P_l \Delta t > S \\ P_l \left(\frac{C}{S}\right) & \text{if } P_l \Delta t < S \end{cases} \qquad (8a)$$

with the grid box mean throughfall given by

$$T = P \left(1 - \frac{C}{S}\right) \exp\left(\frac{-\mu S}{P \Delta t}\right) + P \frac{C}{S} \qquad (8b)$$

The results of 1-D model experiments with both these schemes are also given in table 3. Cases A2-A4 refer to the WJS scheme (Equation 7), whilst cases A5-A7 refer to the new UM scheme (Equation 8). Three situations are considered for each parameterization which relate to the

fractional area covered by a rainstorm. Case A2 and A5 refer to 30% coverage for a convective precipitation event and 100% for large scale rain. Case A3 and A6 refer to 10% and 50% coverage. The cases A4 and A7 use a variable wetted area which is related to the average intensity of rainfall as observed in the Reserva Ducke experiment. As shown in Figure 4, the occurrence of convection shows a marked diurnal variation, so a constant value for the wetted area will ignore this variation and will either produce too weak rainfall over too large an area, or too intense rainfall over too small an area, depending on whether the fixed area is relevant to night time or mid-day convective activity. A more physically based estimate of the fraction can be given by assuming that the diurnal variation can be modelled by incorporating an explicit dependence of the fraction of the grid covered with rain on the grid average rainfall rate. This can be written as

$$\mu = \frac{P}{P_r} \qquad (9)$$

where P is the grid average rainfall rate produced by the model and $P_r$ an observed average rainfall rate as measured in the Reserva Ducke experiment (Lloyd, 1990). From measurements taken at Manaus $P_r$ is given as 5.2 mm $hr^{-1}$.

The parameterization of interception using Equations (7) and (8) suffers from a dependence on the timestep of the model through the incorporation of the rainfall amount in the exponential terms. As a consequence the throughfall decreases when the timestep of the model becomes very small. Typical values for interception loss range from 23% to 17% for a timestep of 5 minutes and 20 minutes respectively with the values changing less rapidly with increasing timestep. To avoid this problem, the calculation of throughfall was performed on amounts adjusted to represent hourly values, and the actual throughfall amounts were adjusted back again each timestep, just before the canopy water budget is drawn up. Another possibility may be to allow $\mu$ to vary with timestep, in which way both the timesteps in Equations (7b) and (8b) will cancel. Both methods ensure that the throughfall percentages remain independent of the timestep and take into account only spatial variability in rainfall.

A further adjustment was made in the models to ensure that evaporation from a wet canopy occurred only from the actual wetted area, with the remainder of the grid freely transpiring. In this case evaporation from wet canopies only takes place in a fraction $\mu$ of the grid box area.

From Table 3 it becomes clear that there is substantial variation in the predicted interception

loss - and corresponding partitioning of evaporation between wet and dry canopy evaporation. Also, slight changes in the predicted rainfall result from changes in the interception parameterization. All parameterizations which allow for a spatial distribution of rainfall show a marked improvement in the prediction of interception and transpiration. Both parameterizations show an decreased interception loss when the area covered with rain is reduced from 30% to 10% (Cases A2,A3 and A5, A6). The Shuttleworth (1988a) parameterization appears to give a slight overestimation of interception loss as compared with the value of 9% ± 3.6 quoted by Lloyd et al. (1988) when a value of $\mu$=0.1 is used (A2). The result obtained by Shuttleworth (1988b), using a calibrated model indicates a loss of 14%. Indeed the model predictions obtained by Lloyd et al (1988) indicate a slight inability to model the observed interception loss, as they predict 12.3% and 11.3% respectively for the two tested models. Both these values are however within measurement error of the observations. The value of 14% (Shuttleworth, 1988b) is probably best looked upon as an upper limit for interception loss in this area. The Unified Model parameterization predicts a 10% interception loss for a value of $\mu$=0.1, which is in agreement with observations (A5). The partitioning of evaporation for this case is shown in Figure 5, which may be compared with Figure 3 to assess the improvement in the evaporation prediction of the model. However, relating the value of $\mu$ to the rainfall rate results in a substantial underestimation of the interception loss for both the Shuttleworth (A4) and UM parameterization (A7). This suggests that the parameterization of $\mu$ gives too small a wetted area in this case. Both parameterizations give remarkably similar results in this case.

To test whether the new parameterization improves the sensitivity of the interception loss to a change in saturation storage capacity, the storage capacity was changed by 50%. This produced a change in interception loss by 15% when using the new UM parameterizatioon with $\mu$=0.1/0.5, which can be compared with a change of 4% for the original scheme, and a change of 25% observed in the stand alone interception modelling by Shuttleworth (1988b) and Lloyd et al. (1988). This indicates that the new parameterization simulates the interception of tropical rainforest on a regional scale much more accurately than the previous parameterization in which no allowance was made for the spatial distribution of rainfall.

*Discussion*

The 1-D model experiments on wet canopy evaporation show the importance of including the effect of spatial inhomogeneity in the parameterization of canopy processes in GCMs. Although some of the previous GCM deforestation experiments predicted the correct total evaporation per day, the partitioning between dry and wet canopy evaporation appeared

to be incorrect. Incorporation of a more physically realistic subgrid parameterization reproduces the correct proportions of dry and wet canopy evaporation, and increases the sensitivity to changes in storage capacity.

The percentage total evaporation is relatively constant in all 1-D model experiments (around 50%). This suggests that the total amount of modelled evaporation is strongly controlled by the radiation input and that changes in the parameterizations of the interception scheme only affect the partitioning between dry and wet canopy evaporation. Although the total amount of moisture thus returned to the atmosphere remains the same, the time scale on which this occurs, is different. Transfer of moisture through transpiration is more gradual and involves longer timescales, through a delay in soil moisture storage, than transfer through interception. This may in turn affect the intensity of runoff or flooding events. From this perspective alone it is important to model the correct balance between wet and dry canopy evaporation.

Shuttleworth (1988b) makes the observation that total evaporation is relatively constant throughout the year at about 110 mm per month. The 1-D model reproduces this phenomenon very clearly, although the almost complete lack of seasonal variation may have been amplified by the slight inability of the model to reproduce the observed rainfall pattern.

The 1-D model experiments show that caution must be exercised in applying subgrid parameterizations straightforwardly, without trying to relate the model results to actual measurements. A similar comment applies to using 1-D vegetation atmosphere models with prescribed atmospheric forcing. With an interception scheme similar to that used in experiments A2-A4, Pitman et al (1990) showed that a change in $\mu$ from 0.5 to 0.1 (for convective storms) gives a 50% reduction in total evaporation. However, important feedback effects are prevented by prescribing the atmospheric forcing. Pitman et al (1990) comment that these may have a strong modifying impact on the sensitivity of their model. The use of a fully 1-D model here demonstrates the importance of such feedbacks on the control of surface fluxes: only a 1% increase in total evaporation is seen between experiments A2 and A3 when $\mu$ is increased from 0.1 to 0.3. A much larger change is found in the partitioning between wet and dry canopy evaporation.

The parameterization in which throughfall is allowed to occur before saturation gave slightly better results than the Shuttleworth (1988b) scheme. In point models of interception loss for tropical rainforest, throughfall before canopy saturation is generally not allowed, and is only observed in very small amounts in the field. Indeed, due to the high intensity of rainfall in the

humid tropics and the relatively short time required to saturate a canopy, the throughfall from unsaturated canopies is usually very small. The somewhat surprisingly good performance of the UM scheme can be explained by realizing that it effectively allows a distribution of the canopy storage over the grid from saturated to unsaturated, while the grid average canopy storage may be unsaturated. By allowing a certain part of the grid to be fully saturated, and the other part not, throughfall from the grid-canopy can still occur, whereas in the Shuttleworth scheme, this would not be allowed until the complete grid average canopy storage is saturated.

The present experiments show the complexity involved in the modelling of the subgrid variability in GCM grids and underline the need to compare GCM model predictions with observations. Such observations could include satellite image analysis of rainfall or cloud cover. It may be possible in the future to relate the fractional coverage to convective cloud amount or cloud liquid water content. The difficulty of this, however, should not be underestimated. As shown in this paper, it may be difficult to assess the relation of model variables like rainfall rate to wetted area in a GCM grid.

It is speculative to predict the consequences of the incorrect partitioning between dry and wet canopy evaporation for the interpretation of past GCM deforestation experiments. The present results indicate that the likely climatic effects such as a weakening of the intensity of the hydrological cycle may have been amplified in the GCM experiments by the overestimation of the interception component of total evaporation. To what amount the compensating mechanism of dry and wet canopy evaporation will disguise this effect can however, only be resolved by new, more realistic GCM experiments which use calibrated land surface models.

**ACKNOWLEDGMENT**

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**LIST OF CAPTIONS**

Figure 1. *Schematic diagram of the land surface scheme as used in UKMO GCM and in the original version of the 1-D model (Redrawn from Warrilow and Buckley, 1989). Rsw, Rlw short and longwave radiation, α albedo, Rc and Rl convective and large scale precipitation, Ew and Ed wet and dry canopy evaporation, H sensible heat flux, $T_i$ temperature of the $i^{th}$ soil layer, D rooting depth, Θ soil moisture content, Qs and Qg surface runoff and gravitational drainage, C canopy water content and $T_f$ throughfall.*

Figure 2. *Comparison of observed,(——) (average of 1971-1985) and predicted (— —) monthly rainfall (mm).*

Figure 3. *Partitioning of total evaporation into interception and transpiration for the original parameterization. Modelled rainfall is also shown (—).*

Figure 4. *Comparison of observed (——) and predicted (.....) frequency distribution of rainfall.*

Figure 5. *Partitioning of total evaporation into interception and transpiration for the UM parameterization with μ=0.1/0.5 (run A5). Modelled rainfall is also shown (—).*

**TABLE 1a. 1-D Model forcing data, for Manaus, Brazil (3°S, 60°W), January**

T is average monthly temperature, σT standard deviation of temperature, D average monthly dewpoint depression, δT and δD gradient of temperature and dewpoint depression from the edge to the 1-D model area centre, $V_n$ and $V_p$ average horizontal wind velocity components, ω vertical p-velocity and σW the standard deviation of the vertical p-velocity.

| Level | 11 | 10 | 9 | 8 | 7 | 6 | 5 | 4 | 3 | 2 | 1 |
|---|---|---|---|---|---|---|---|---|---|---|---|
| T (°K) | 299.0 | 296.0 | 291.0 | 283.5 | 273.0 | 260.0 | 245.0 | 229.0 | 213.0 | 194.0 | 213.0 |
| σT (°K) | 2.0 | 2.0 | 2.0 | 2.0 | 1.7 | 1.5 | 1.5 | 1.5 | 1.5 | 2.2 | 3.0 |
| D (°K) | 4.0 | 4.0 | 5.5 | 6.5 | 12.5 | 8.0 | 8.0 | 6.0 | 8.0 | 10.0 | 30.0 |
| δT (°K) | 1 | 1 | 1 | 1 | 1 | 1 | 1 | 1 | 1 | 1 | 1 ($*10^{-4}$) |
| δD (°K) | 1 | 1 | -60 | -150 | -270 | -430 | -570 | 1 | 1 | 1 | 1 ($*10^{-4}$) |
| Vn (m/s) | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| Vp (m/s) | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| σVn (m/s) | 2 | 2.2 | 2.5 | 3.1 | 3.6 | 4.2 | 4.8 | 5.5 | 6.1 | 6.7 | 5.7 ($*10^{-5}$) |
| ω (mb/s) | -75 | -150 | -300 | -570 | -600 | -570 | -465 | -330 | -165 | 75 | 105 ($*10^{-7}$) |
| σω (mb/s) | 1 | 2 | 4 | 7.6 | 8.0 | 7.6 | 6.2 | 4.4 | 2.2 | 1.0 | 1.4 ($*10^{-4}$) |

**TABLE 1b. 1-D Model forcing data, for Manaus, Brazil (3°S, 60°W), July**

See Table 1a for explanation of symbols.

| Level | 11 | 10 | 9 | 8 | 7 | 6 | 5 | 4 | 3 | 2 | 1 |
|---|---|---|---|---|---|---|---|---|---|---|---|
| T (°K) | 297.0 | 294.5 | 289.0 | 282.0 | 271.5 | 258.0 | 243.0 | 228.0 | 208.0 | 197.0 | 218.0 |
| σT (°K) | 2.0 | 2.0 | 2.0 | 1.3 | 1.3 | 1.6 | 1.9 | 1.5 | 1.5 | 2.5 | 2.5 |
| D (°K) | 1.0 | 3.0 | 4.5 | 10.0 | 12.0 | 12.5 | 14.0 | 15.0 | 24.0 | 14.0 | 35.0 |
| δT (°K) | 27 | 14 | 4.5 | 4.5 | 4.5 | 4.5 | 4.5 | 2.4 | 0.63 | 0.63 | 0.63 ($*10^{-4}$) |
| δD (°K) | -23 | -61 | -120 | -200 | -310 | -440 | -620 | 2.4 | 0.63 | 0.63 | 0.63 ($*10^{-4}$) |
| Vn (m/s) | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| Vp (m/s) | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 |
| σVn (m/s) | 2 | 2.2 | 2.6 | 3.1 | 3.6 | 4.4 | 5.0 | 5.6 | 6.0 | 5.5 | 5.0 ($*10^{-5}$) |
| ω (mb/s) | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 | 0 ($*10^{-7}$) |
| σω (mb/s) | 0 | 1.0 | 3.0 | 4.0 | 4.2 | 4.2 | 4.2 | 3.0 | 1.0 | 1.0 | 1.0 ($*10^{-4}$) |

**TABLE 2. Parameters used in the land surface scheme for tropical forest**

| | |
|---|---|
| *surface albedo(%)* | *12.25* |
| *surface resistance (s $m^{-1}$)* | *115* |
| *saturation storage capacity (mm)* | *0.7* |
| *fractional vegetation cover* | *0.9* |
| *rooting depth(m)* | *1.5* |
| *surface roughness (m)* | *2.0* |

**TABLE 3. 1-D Model interception and transpiration results.**

| MODEL VERSION | "Observed" | A1 | A2 | A3 | A4 | A5 | A6 | A7 |
|---|---|---|---|---|---|---|---|---|
| PRECIPITATION (mm) | 2593 | 2418 | 2347 | 2367 | 2332 | 2215 | 2359 | 2332 |
| INTERCEPTION (mm) | 373 | 766 | 316 | 457 | 127 | 222 | 406 | 125 |
| Percentage | 14 | 32 | 14 | 19 | 6 | 10 | 17 | 5 |
| TRANSPIRATION (mm) | 1020 | 512 | 873 | 760 | 1036 | 950 | 799 | 1038 |
| Percentage | 39 | 21 | 37 | 32 | 44 | 43 | 34 | 45 |

*Models:*

*A1 'Old' GCM Parameterization*

*A2 WJS with μ=0.1/0.5*

*A3 WJS with μ=0.3/1.0*

*A4 WJS with variable wetted area*

*A5 'New' Unified model Parameterization, μ=0.1/0.5*

*A6 'New' Unified model Parameterization, μ=0.3/1.0*

*A7 'New' Unified model Parameterization with variable wetted area (μ=0.1 or μ=0.3 applies to convective rainfall, μ=0.5,μ=1 applies to large scale rainfall)*

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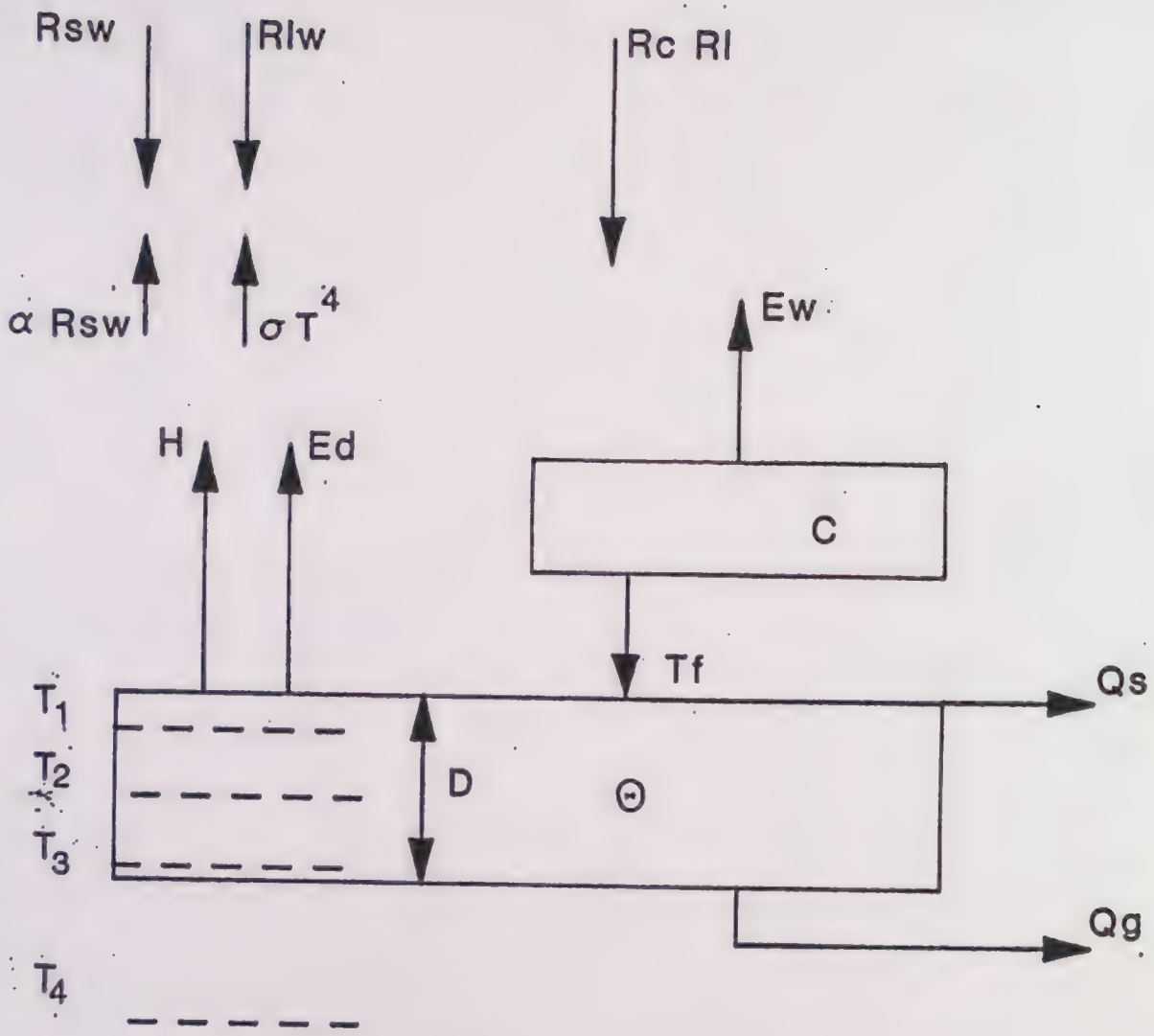


Figure 1

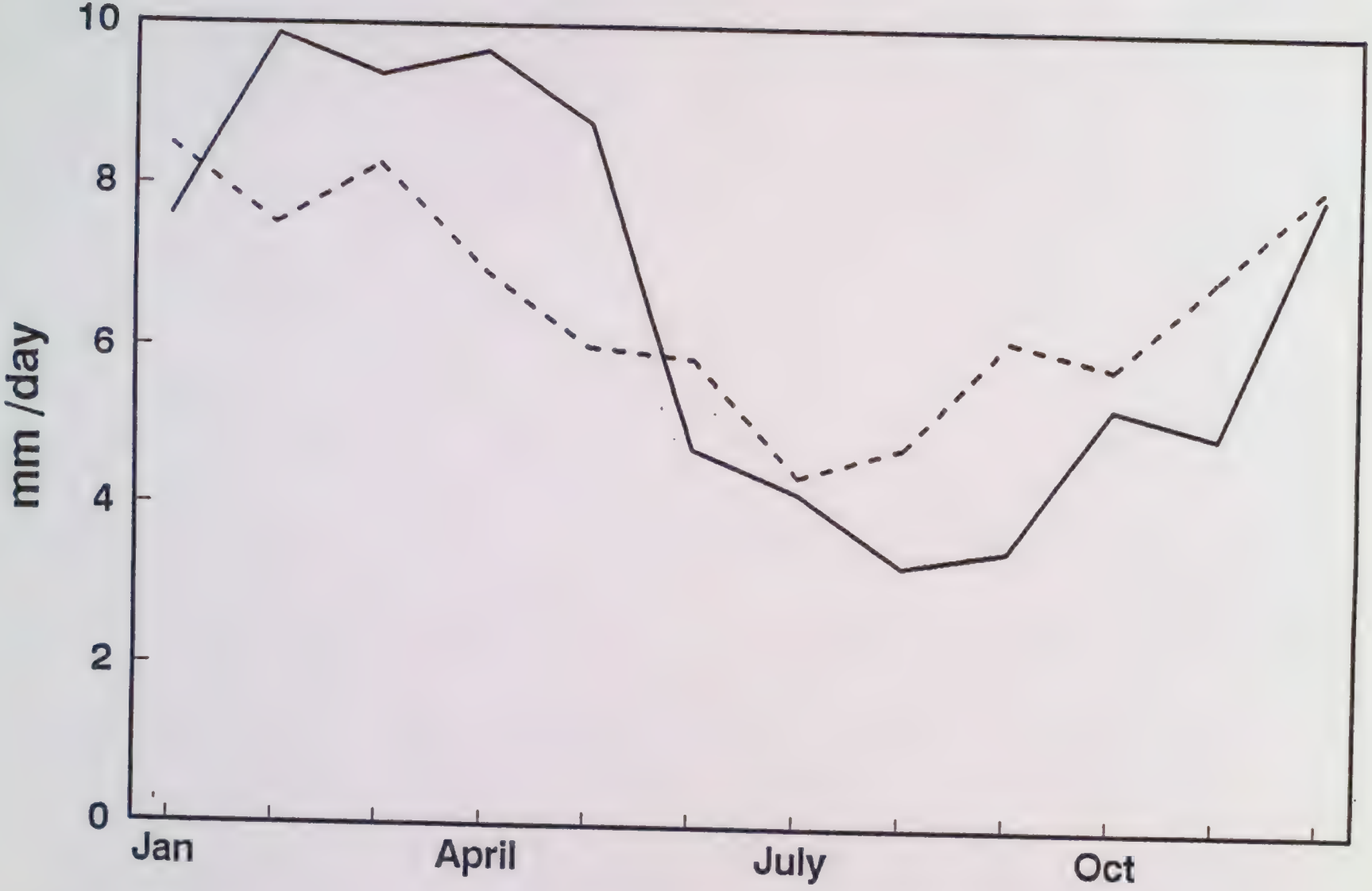


Figure 2

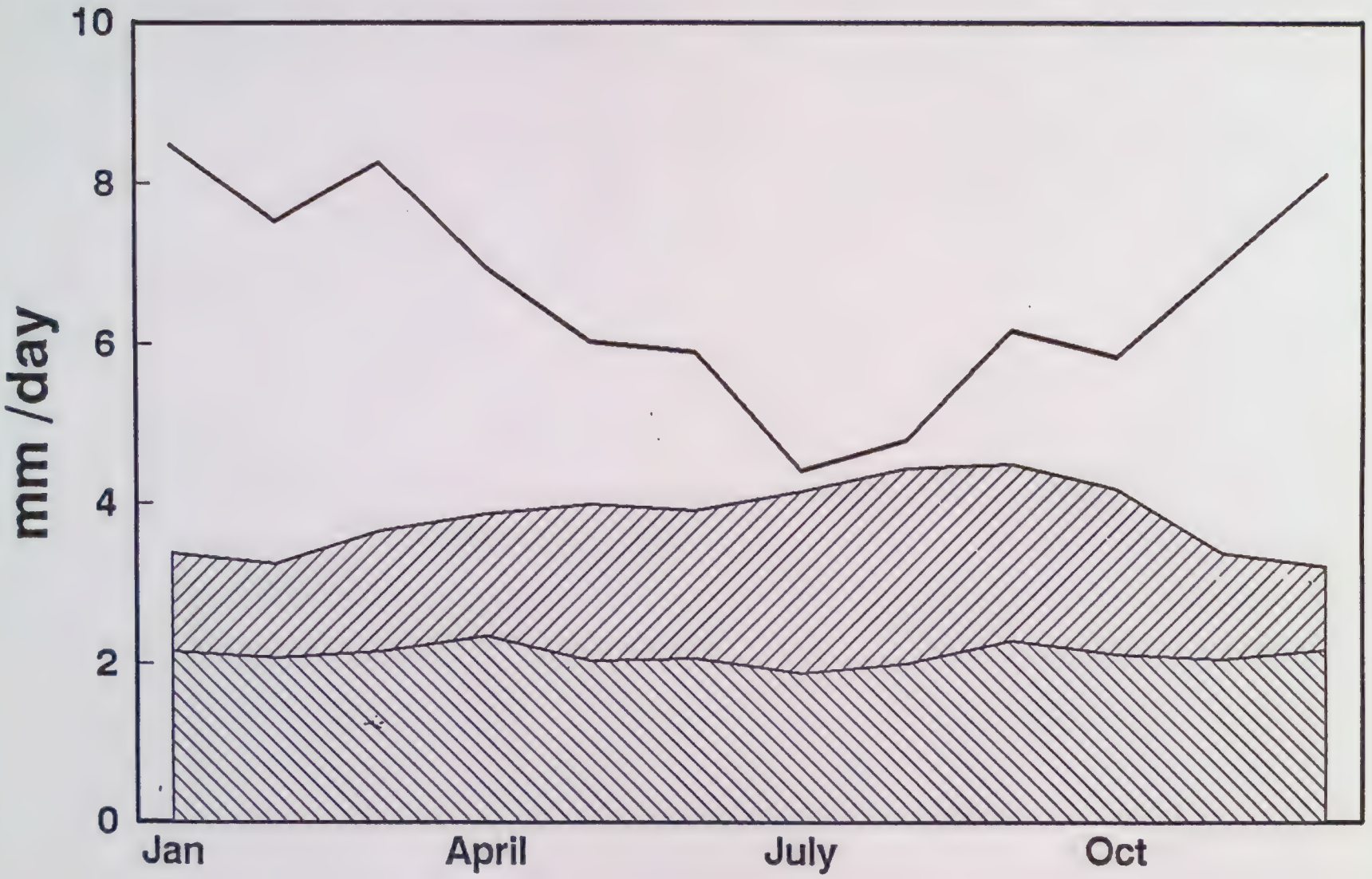


Figure 3

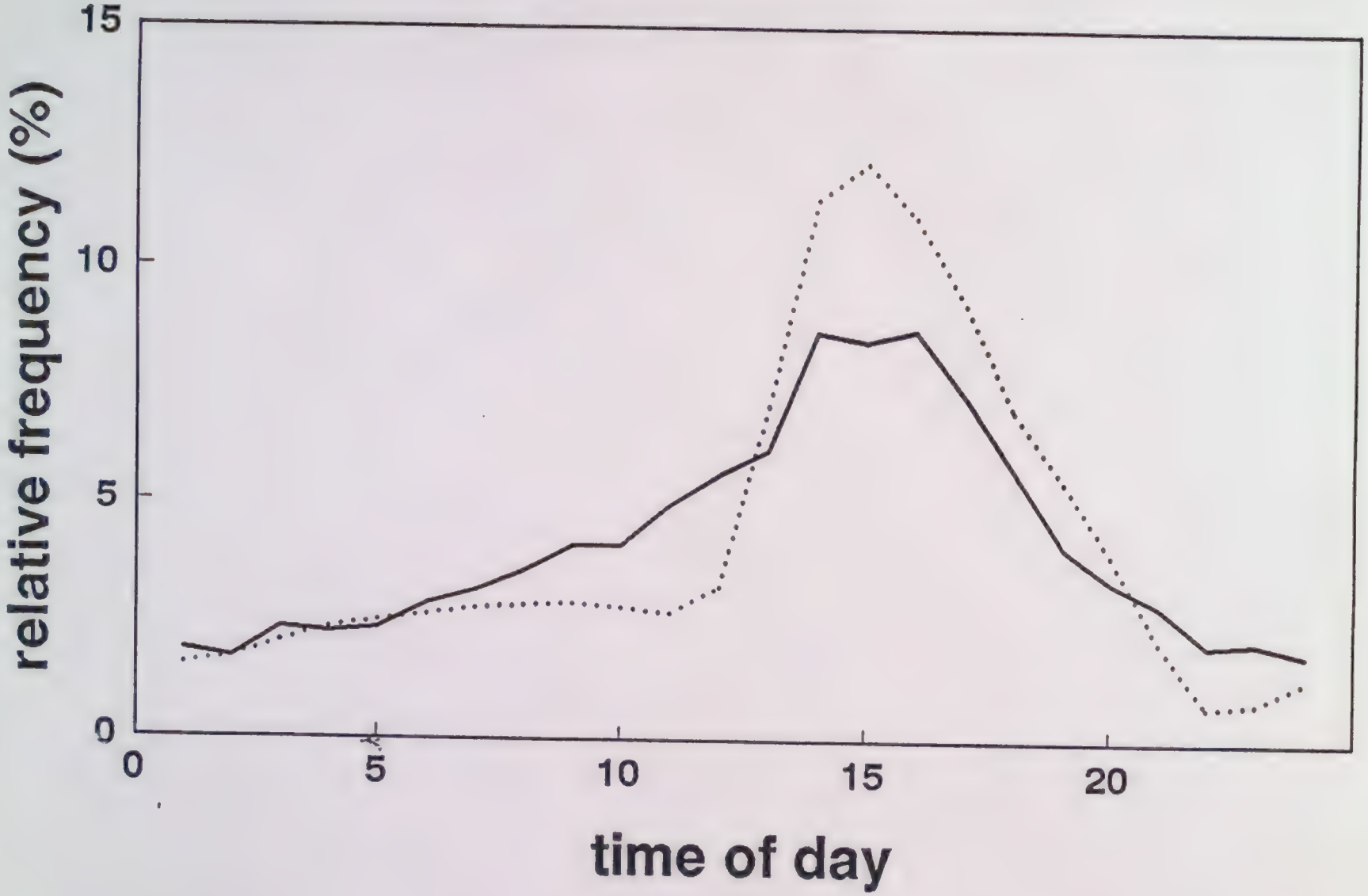


Figure 4

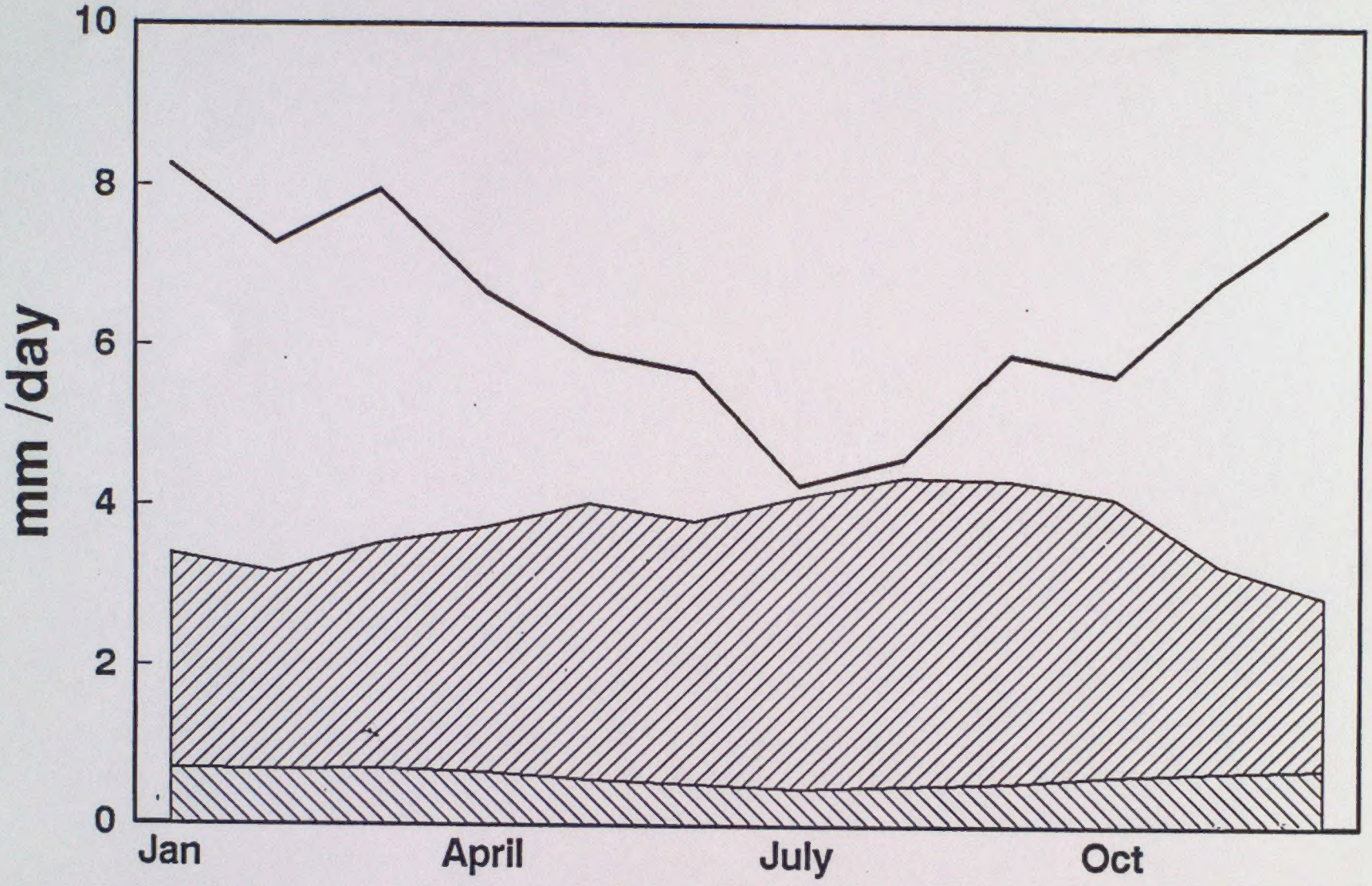


Figure 5

CLIMATE RESEARCH TECHNICAL NOTES

| | | |
|---|---|---|
| CRTN 1 | Oct 1990 | Estimates of the sensitivity of climate to vegetation changes using the Penman-Monteith equation.<br>P R Rowntree |
| CRTN 2 | Oct 1990 | An ocean general circulation model of the Indian Ocean for hindcasting studies.<br>D J Carrington |
| CRTN 3 | Oct 1990 | Simulation of the tropical diurnal cycle in a climate model.<br>D P Rowell |
| CRTN 4 | Oct 1990 | Low frequency variability of the oceans.<br>C K Folland, A Colman, D E Parker and A Bevan |
| CRTN 5 | Dec 1990 | A comparison of 11-level General Circulation Model Simulations with observations in the East Sahel.<br>K Maskell |
| CRTN 6 | Dec 1990 | Climate Change Prediction.<br>J F B Mitchell and Qing-cun Zeng |
| CRTN 7 | Jan 1991 | Deforestation of Amazonia - modelling the effects of albedo change.<br>M F Mylne and P R Rowntree |
| CRTN 8 | Jan 1991 | The role of observations in climate prediction and research.<br>D J Carson |
| CRTN 9 | Mar 1991 | The greenhouse effect and its likely consequences for climate change.<br>D J Carson |
| CRTN 10 | Apr 1991 | Use of wind stresses from operational N.W.P. models to force an O.G.C.M. of the Indian Ocean.<br>D J Carrington |
| CRTN 11 | Jun 1991 | A new daily Central England Temperature series, 1772-1991.<br>D E Parker, T P Legg and C K Folland |
| CRTN 12 | Jul 1991 | Causes and predictability of Sahel rainfall variability.<br>D P Rowell, C K Folland, K Maskell, J A Owen, M N Ward |
| CRTN 13 | Jul 1991 | Modelling changes in climate due to enhanced $CO_2$, the role of atmospheric dynamics, cloud and moisture.<br>C A Senior, J F B Mitchell, H Le Treut and Z-X Li |

CLIMATE RESEARCH TECHNICAL NOTES

| | | |
|---|---|---|
| CRTN 14 | Sep 1991 | Sea temperature bucket models used to correct historical SST data in the Meteorological Office.<br>C K Folland |
| CRTN 15 | Aug 1991 | Modelling climate change, and some potential effects on agriculture in the U.K.<br>P R Rowntree, B A Callander and J Cochrane |
| CRTN 16 | Aug 1991 | The Boreal Forests and Climate<br>G Thomas and P R Rowntree |
| CRTN 17 | Aug 1991 | Development of a Stratosphere-Troposphere Data Assimilation System.<br>R Swinbank. |
| CRTN 18 | Sep 1991 | A study of asynchronous coupling using a simple climate model.<br>M K Davey. |
| CRTN 19 | Sep 1991 | The Oceanic Carbon Cycle.<br>N K Taylor |
| CRTN 20 | Nov 1991 | Worldwide ocean-atmosphere surface fields in Sahel wet and dry years using provisionally corrected surface wind data.<br>M N Ward |
| CRTN 21 | Dec 1991 | Coupled tropical ocean global atmosphere models at the UKMO.<br>M K Davey, C Gordon, S Ineson and S Lawrence |
| CRTN 22 | Dec 1991 | Empirical parameterisation of tropical ocean-atmosphere coupling.<br>M Allen, M K Davey, D L T Anderson and P D Killworth |
| CRTN 23 | Feb 1992 | The temporal evolution of Equatorial Currents in the Indian Ocean<br>D L T Anderson, D J Carrington, R A Corry and C Gordon |
| CRTN 24 | Feb 1992 | Stratospheric analyses provided by the U.K. Meteorological Office<br>M J Bailey, A O'Neill and V D Pope |
| CRTN 25 | Feb 1992 | Modelling interannual variability in the Indian Ocean using momentum fluxes from the UKMO and ECMWF operational weather analyses.<br>D L T Anderson and D J Carrington |

CLIMATE RESEARCH TECHNICAL NOTES

| | | |
|---|---|---|
| CRTN 26 | Mar 1992 | A GCM simulation of the impact of Amazonian deforestation on climate using an improved canopy representation<br>J Lean and P R Rowntree |
| CRTN 27 | Mar 1992 | The parameterization of rainfall interception in GCMs<br>A J Dolman and D Gregory |